Neoproterozoic Glaciation: Reconciling Low Paleolatitudes and the Geologic Record

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Paleomagnetic data for Neoproterozoic glacial deposits in South Australia and elsewhere verify glaciomarine deposition near the paleoequator. Tidal rhythmites from such deposits in South Australia display symmetrical ripples indicating decades of continuous wave activity, and also record the annual oscillation of sea level. In low and moderate latitudes the annual oscillation of sea level results mostly from seasonal changes in heat content of the sea, indicating extensive and long-lived open seas in low latitudes during Neoproterozoic glaciations. Neoproterozoic periglacial sand wedges 3+ m deep, marking polygons 10–30 m across, are closely comparable to periglacial wedges in present high latitudes and imply large (~40°C) seasonal changes of mean monthly temperature near the paleoequator. Periglacial wedges did not form at high elevations on the Pleistocene equator where temperatures were well below 0°C throughout the year and temperature fluctuations were mainly diurnal, which militates against diurnal fluctuations as the cause of the Neoproterozoic wedges. An extremely large (50%) octupole component of the geomagnetic field is required to make true moderate latitudes appear paleoequatorial, whereas Proterozoic paleomagnetic data suggest a maximum octupole component of ≤30%. A snowball or slushball Earth is difficult to reconcile with open seas and large seasonal temperature-changes in low paleolatitudes. A Proterozoic high obliquity (>54°) resulting from the Moon-producing single giant impact may explain glaciation and strong seasonality on the equator, but a mechanism is required to subsequently reduce the obliquity. At present the Neoproterozoic paleomagnetic and glacial records cannot be reconciled satisfactorily, demanding further wide-ranging research.

1. INTRODUCTION

Neoproterozoic glaciogenic rocks are widely distributed and include thick marine tillites, mudstone-with-dropstones facies, glaciofluvial outwash, striated pavements, and permafrost regoliths. Although such glacial features are like those of the Phanerozoic, the Neoproterozoic successions have generated much controversy because paleomagnetic data indicate some formed near the paleoequator, carbonates are associated with many glacial deposits, and iron-formations occur within several successions.

The three possible scenarios that seem most able to explain the enigma of low-paleolatitude glaciation are (1) a gross

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failure of the geocentric axial dipole hypothesis to describe the Proterozoic geomagnetic field [*Kent and Smethurst*, 1998], (2) a 'snowball Earth' [*Kirschvink*, 1992; *Hoffman and Schrag*, 2002], and (3) a high obliquity of the ecliptic [*Williams*, 1975, 1993]. Here we examine how these three disparate scenarios may be reconciled with key paleomagnetic and geologic observations.

2. PROTEROZOIC LOW-PALEOLATITUDE GLACIATION

The early suggestion that Neoproterozoic glaciation reached low paleolatitudes [*Harland*, 1964] had few adherents for many years, despite some supporting studies [e.g., *McWilliams and McElhinny*, 1980]. However, there is now irrefutable evidence that Neoproterozoic glaciomarine deposition occurred near the paleoequator. The best evidence for low paleolatitudes currently comes from paleomagnetic studies of the Marinoan (~600 Ma) Elatina Formation in South Australia [*Embleton and Williams*, 1986; *Schmidt et al.*, 1991; *Schmidt and Williams*, 1995; *Sohl et al.*, 1999]. Paleomagnetic data for drill-core from Western Australia support a low paleolatitude for Marinoan glaciation and imply a low paleolatitude also for Sturtian (~750–700 Ma) glaciation [*Pisarevsky et al.*, 2001]. Low-inclination paleomagnetic data for the Rapitan glacial deposits in northwestern Canada [*Park*, 1997], which have been correlated with glacial deposits and associated 685 Ma volcanic rocks in Idaho [*Lund et al.*, 2003], also imply

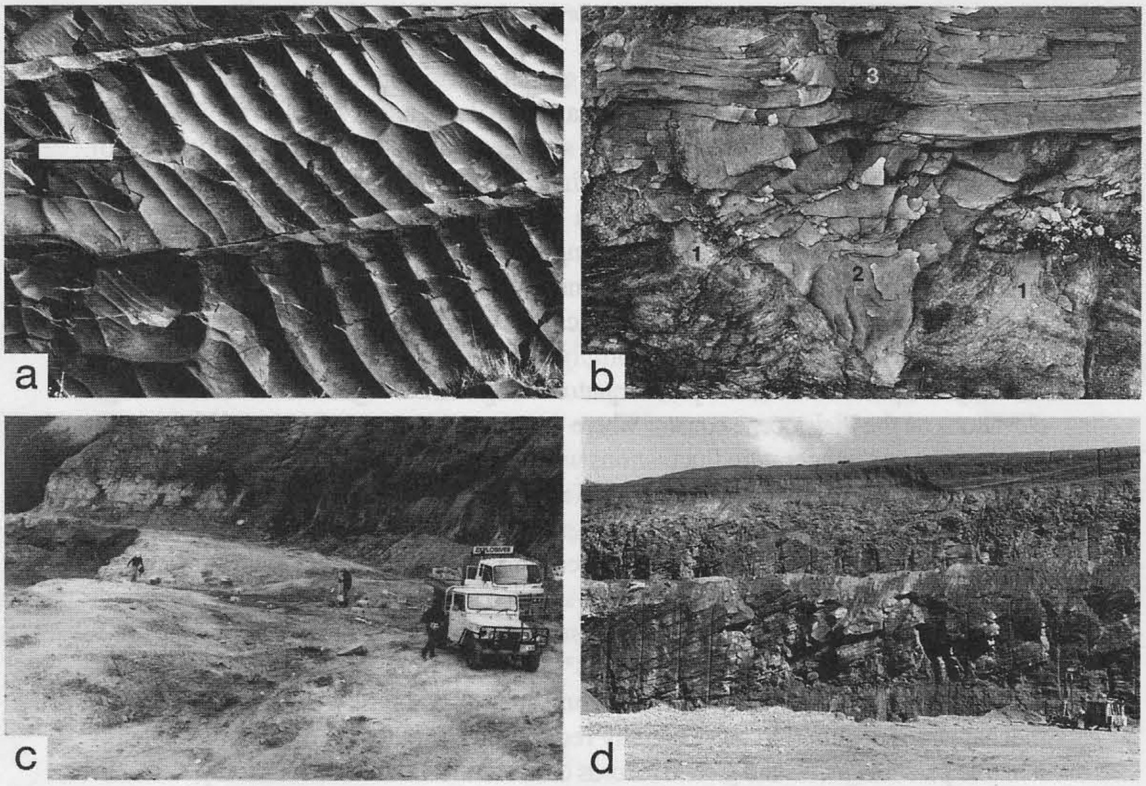

Figure 1. (a) Symmetrical ripples on a bed surface of the Elatina Formation (Marinoan), Warren Gorge, South Australia. Scale 15 cm. (b) Marinoan periglacial sand wedges, Cattle Grid mine, South Australia. The large wedge (2) is 3 m deep, comprises steeply dipping laminae of pebbly coarse sandstone, and is developed in a permafrost regolith of in situ frost-shattered Mesoproterozoic quartzite (Cattle Grid Breccia). Two deformed sand wedges of an earlier generation (1) occur within the breccia, and a third generation wedge (3) occurs in the upper part of the large wedge and the overlying periglacial–eolian Whyalla Sandstone (Marinoan). Upturning of material next to the wedges records summer expansion of the permafrost. Closely comparable large-scale sand wedges are forming today in Antarctica under a strongly seasonal climate. (c) A mine bench in the Cattle Grid mine excavated to the top of the Cattle Grid Breccia, showing hummocks >10 m across; the depressions between the hummocks (by figure at left) mark the position of sand wedges. The vertical mine face in the background is of Whyalla Sandstone. Photograph by courtesy of J. L. Curtis. (d) Cross-strata in the periglacial–eolian Whyalla Sandstone, Cattle Grid mine; two cross-bed sets each ~7 m thick overlie basal eolian low-angle strata (behind vehicle). The section is capped by Quaternary sand.

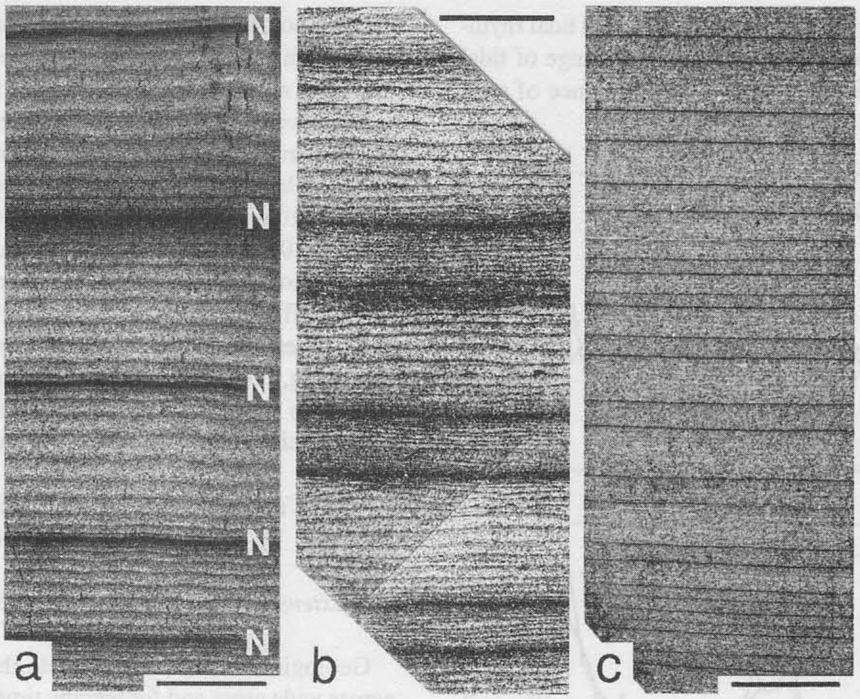

Figure 2. Neoproterozoic tidal rhythmites, South Australia. Neap-tidal mudstone bands (N), where some abbreviation of synodic fortnightly neap–spring cycles has occurred, appear darker than sandy and silty laminae. Scale bars 1 cm. (a) Elatina Formation (~600 Ma), Pichi Richi Pass, showing four neap–spring cycles of graded diurnal laminae. (b) Chambers Bluff Tillite (~750–700 Ma), showing six neap–spring cycles of semidiurnal and diurnal laminae. Successive relatively thick and thin neap–spring cycles reflect the 'monthly inequality' of spring-tidal height (range) and current speed due to the elliptical lunar orbit. (c) Pualco Tillite (~750–700 Ma), showing 25 distal neap–spring cycles.

low-latitude glaciation. Paleomagnetic results for 608 Ma dikes in Sweden place Baltica on the paleoequator during the Varanger glaciations [*Eneroth and Svenningsen*, 2004]. Furthermore, paleomagnetic data for Paleoproterozoic volcanic rocks in South Africa [*Evans et al.*, 1997] and sedimentary rocks and mafic dikes in Canada [*Williams and Schmidt*, 1997; *Schmidt and Williams*, 1999] suggest low paleolatitudes for associated 2.3–2.2 Ga glaciogenic rocks.

The arguments of *Embleton and Williams* [1986] for the early timing of remanence of the Elatina Formation have been fully vindicated. Mineragraphic identification of detrital high-temperature forms of titano-hematite, coupled with the low metamorphic grade (diagenetic temperatures of <135–160°C) and discrete unblocking temperatures near 680°C (the hematite Curie temperature), effectively ruled out chemical or thermal remagnetization. However, perhaps the strongest evidence that the Elatina Formation acquired its magnetization early comes from positive fold tests on soft-sediment folds that formed by gravity sliding on a tidal delta [*Williams*, 1996]; such gravity-flow structures are typical of deltaic deposits [*Myrow and Hiscott*, 1991]. Three separate fold tests have been executed and each has proved positive [*Sumner et al.*, 1987; *Schmidt et al.*, 1991; *Schmidt and Williams*, 1995], confirming detrital remanent magnetization.

None of the above studies, however, answered growing objections that the Elatina paleopole was a virtual geomagnetic pole, having sampled the field for <100 years. Subsequently, two independent regional paleomagnetic studies that provided concordant paleomagnetic directions ascribed to early chemical remanent magnetization removed those doubts [*Schmidt and Williams*, 1995; *Sohl et al.*, 1999]. The combined data yield a paleopole at 41.6°S, 359.6°E (dp = 3.0°, dm = 5.9°) and a paleolatitude of 7.9 ± 3.0° [*Schmidt*, 2001]. Neither study provided a complete magnetostratigraphy, although *Schmidt and Williams* [1995] noted a rough magnetostratigraphy where polarity changes from reverse to normal up-section and *Sohl et al.* [1999] identified at least three magnetic polarity intervals. *Schmidt and Williams* [1995] also recorded magnetization reversals within specimens suggesting that, for some of the time at least, the magnetization and the reversal processes shared a similar time-constant that resulted in many zones of mixed polarity.

Tidal deposits in Marinoan and Sturtian glaciogenic successions in South Australia display herringbone cross-bedding, marking successive ebb and flood tides, and tidal rhythmites (cyclic tidal deposits) recording a wide range of tidal cycles [*Williams*, 1989, 1991, 2000]. The presence of such deposits confirms glaciomarine deposition. Cold climate near sea level in low paleolatitudes is all the more puzzling because the faster rotation of the Proterozoic Earth [*Williams*, 2000] would have caused less efficient poleward transport of heat, resulting in slightly warmer equatorial regions and substantially cooler poles [*Kuhn et al.*, 1989]. Unequivocal evidence for Neoproterozoic high-paleolatitude (>60°) deposits is lacking [*Evans*, 2000]. Currently it is unclear whether Neoproterozoic glaciomarine deposition preferentially in low paleolatitudes reflects true cold-climate distribution, a lack of landmasses in high paleolatitudes, or insufficient data. It is intriguing, nonetheless, that the enigma of Neoproterozoic glaciation near sea level seemingly favoring low paleolatitudes persists after four decades of research.

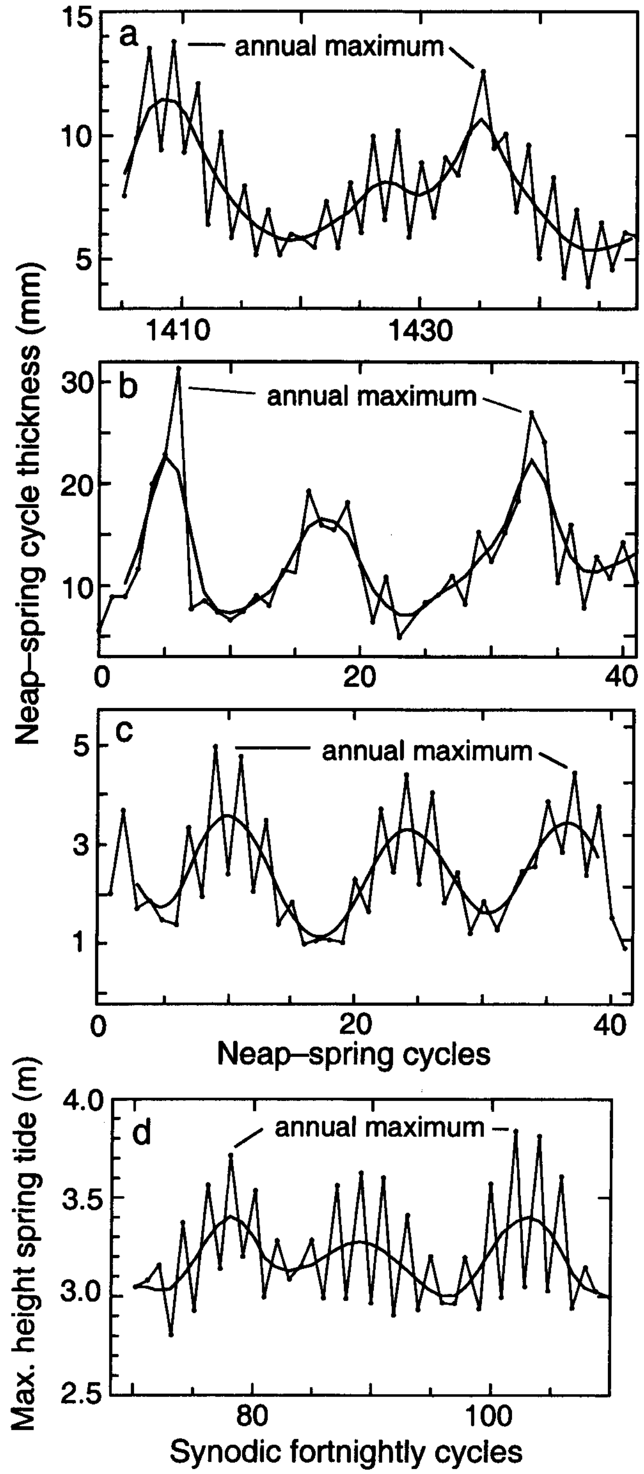

Figure 3. Neap–spring cycle thickness for Neoproterozoic tidal rhythmites (the cycle numbers increase progressing up the stratigraphic successions), and modern tidal data. (a) Part of the 60-year Elatina record. The sawtooth pattern reflects the 'monthly inequality'. (b) Chambers Bluff Tillite; depositional processes have obscured the monthly inequality in some places. (c) Pualco Tillite. (d) Maximum height of the synodic fortnightly tidal cycle at Townsville, Queensland, from October 19 1968 to June 3 1970. The four smoothed curves (five-point filter weighted 1,4,6,4,1) highlight the non-tidal annual oscillation of sea level that defines the solar year. The second-order peaks in the curves mark the semiannual tidal cycle, which is in phase with the annual oscillations and whose amplitude is modulated by the lunar nodal cycle [*Williams*, 2000]. Data for Townsville provided by the Beach Protection Authority, Queensland Department of Transport.

3. KEY FEATURES OF THE NEOPROTEROZOIC GLACIAL ENVIRONMENT

3.1. Extensive and Long-lived Open Seas

Geologic observations indicate that seas were unfrozen across wide areas and for lengthy time intervals during Marinoan and Sturtian glaciations in South Australia.

Tidal rhythmites in the Elatina Formation display wave-generated symmetrical ripples (Figure 1a) and interference ripples throughout a 20 m thick succession that records decades of deposition with continuous wave activity and open seas [*Williams*, 1996].

Tidal rhythmites elsewhere in the Elatina Formation (Figure 2a) record a wide range of paleotidal periods and the annual (or seasonal) oscillation of sea level through regular variation in the thickness of neap–spring (synodic fortnightly) cycles (Figure 3a), which is a proxy for tidal height and range [*Williams*, 2000]. The conspicuous annual oscillation averages 26.2 ± 0.9 (1 σ) neap–spring cycles and occurred continuously for at least 60 years throughout Elatina rhythmite deposition. Tidal rhythmites from the Sturtian Chambers Bluff Tillite (Figure 2b) in northern South Australia also record a clear annual oscillation (Figure 3b). Distal tidal rhythmites in the Sturtian Pualco Tillite (Figure 2c) 700 km to the southeast display a less conspicuous but discernible annual oscillation (Figure 3c). The Sturtian oscillations contain 26–28 neap–spring cycles, consistent with an increasing number of months per year going back in time.

The annual oscillation of sea level is of non-tidal origin, resulting from a complex interaction of physical processes including changes in atmospheric pressure, winds, and water temperature [*Komar and Enfield*, 1987]. In most places the highest sea level occurs in the autumn, with the oscillations in the northern and southern hemispheres being out of phase. The period of the oscillation at a particular site thus reflects the solar year. The oscillation is recorded by the maximum height of spring tides at Townsville, Queensland (Figure 3d) [*Williams*, 2000]. *Pattullo* [1966] concluded that the changing heat content of the sea explains most of the observed annual variation in sea level between 45°N and 45°S. At Bermuda the annual oscillation is 'primarily thermal', with winds being of minor importance [*Wunsch*, 1972, p. 32]. The strong annual oscillation at Balboa is directly related to sea surface temperature [*Roden*, 1963]. Furthermore, *Mellor and Ezer* [1995] found that the annual variation of sea level for the Atlantic Ocean between 66°N and 66°S approximates the heating–cooling cycle of each hemisphere. If the seas had been frozen-over during Neoproterozoic glaciations (see Section 4.2), the annual oscillation of sea level could not have occurred because the ice cover would have isolated the sea from seasonal changes of temperature and winds.

The geologic findings of extensive and long-lived open seas are supported by positive $\delta^{13}C_{carb}$ values for carbonates within and directly above Neoproterozoic glaciogenic successions in Namibia, Australia and North America, which imply a very normal, open ocean [*Kennedy et al.*, 2001]. Moreover, *McMechan* [2000] and *Condon et al.* [2002] argued that the thick, widespread mudstone-with-dropstones facies in numerous Neoproterozoic glaciogenic successions attest to temperate glacial conditions during long intervals with voluminous sediment-laden meltwater plumes and icebergs calving into open seas. Their views, while contentious [*Donnadieu et al.*, 2003], are consistent with the present findings.

3.2. Strongly Seasonal Climate in Low Paleolatitudes

Large seasonal changes of temperature are implied by spectacular Marinoan periglacial sand wedges that formed in a then coastal area in South Australia (Figure 1b,c). Four generations of sand wedges as much as 3+ m deep and marking polygons 10–30 m across occur in a permafrost regolith of in situ frost-shattered quartzite averaging 5 m in thickness and in overlying periglacial eolian sandstone [*Williams and Tonkin*, 1985; *Williams*, 1986, 1994a, 1998]. Associated meter-scale features include anticlines and tepee-like structures, some cut by reverse faults, indicating frost heaving and frost thrusting; truncated earth mounds; frost-heaved boulders; diapiric breccia injections; periglacial involutions; and a steep-sided channel filled with varve-like beds. The Marinoan wedges and the range and large dimensions of associated features are very like periglacial structures found in modern polar regions [*Péwé*, 1959; *Black*, 1982; *Washburn*, 1980; *Karte*, 1983]. Neoproterozoic periglacial wedges occur also in Mauritania [*Deynoux*, 1982], Scotland, Spitsbergen, Norway [see summary in *Williams*, 1986], and Greenland [*Moncrieff and Hambrey*, 1990].

Sand wedges and ice wedges are confined to periglacial regions marked by a *strongly seasonal* climate [*Washburn*, 1980; *Karte*, 1983]. Sand wedges are best developed in the dry valleys of Antarctica where the mean monthly air temperature (MMAT) ranges from –35°C or lower in mid-winter to +4°C or lower in midsummer. Wedges show vertical lamination and in plan they define polygons ~10–30 m across. It is widely agreed that the wedges develop from thermal contraction cracks ~1–5 mm wide and several meters deep that form in the upper part of permafrost with rapid drops of temperature during repeated severe winters. The permafrost may fracture with sharp, explosive-like noises. Seasonal temperature changes may reach depths of 15 m or more, whereas diurnal fluctuations of temperature may affect the uppermost permafrost to a depth of 1 m at the most [*Embleton and King*, 1975]. Ice wedges occur in relatively humid periglacial areas where water freezes in the cracks, and sand wedges mark drier periglacial areas where the cracks are filled by drifting sand. Measurements across sand wedges in Antarctica over two decades indicated mean growth rates of up to 1 mm yr^{-1} [*Black*, 1982], and estimated ages of ice wedges in Alaska based on measured growth rates of 1–3 mm yr^{-1} were verified by radiocarbon dating [*Black*, 1952, 1982]. These observations confirm that periglacial wedges are actively forming in high latitudes under strongly seasonal climates.

Periglacial sand wedges are viewed as reliable indicators of past climate [*Washburn*, 1980; *Karte*, 1983]. The Marinoan sand wedges occur in quartzose materials and the polygonal cracking can be ascribed with confidence to thermal contraction. These wedges imply a frigid, strongly seasonal climate (MMAT range of ~40°C), which occurred in a coastal area near the paleoequator. In present low latitudes (≤10°), by contrast, the MMAT range near sea level is <2°C and the mean diurnal air temperature range is <10°C [*Müller*, 1982].

Maloof et al. [2002] conducted modeling that suggested diurnal temperature fluctuations during a snowball Earth event may produce thermal contraction cracks 1+ m deep, concluding that the Marinoan sand-wedge polygons 3+ m deep and 10–30 m across may record diurnal temperature fluctuations. However, only miniature periglacial forms occur on presently glaciated equatorial mountains, where temperature fluctuations of a few °C are largely diurnal [*Zeuner*, 1949; *Hastenrath*, 1981]. In Tanzania and Kenya, mud polygons and sorted stone polygons 15–20 cm in diameter and 3–4 cm deep occur at >4500 m on Mount Kilimanjaro (5899 m) at 3°S and at 4200–4300 m on Mount Kenya (5199 m) at 0.1°S [*Hastenrath*, 1973]. Such small, shallow periglacial forms also occur at 4300–5000 m in the Ecuadorian Andes, which reach 6310 m at 1.2°S [*Hastenrath*, 1981]. All these periglacial forms occur where the mean annual air temperature (MAAT) is near or below 0°C, and taking a mean lapse rate of 0.65°C/100 m gives MAATs as low as –7°C to –10°C at highest elevations. Mean temperatures on equatorial mountains were 4°C to 6°C lower during late Pleistocene stades [*Bonnefille et al.*, 1990; *Clapperton*, 1993], implying former MAATs as low as –11°C to –16°C and temperatures well below 0°C throughout the year. Such figures are below the required MAATs of 0°C for the occurrence of permafrost and –2°C to –10°C for permafrost cracking responsible for large-scale ice- and sand-wedge polygons [*Washburn*, 1980]. However, late Pleistocene large-scale polygonal wedges are absent on equatorial mountains despite the maintenance of requisite low temperatures for several millennia and the presence of suitable low-gradient surfaces. Miniature periglacial forms characterize high elevations in equatorial latitudes evidently because such regions experience only diurnal freeze–thaw cycles, which are of much smaller amplitude and duration than the seasonal freeze–thaw cycles of high latitudes.

The above observations argue against diurnal temperature fluctuations as the cause of Neoproterozoic periglacial sand-wedge polygons. Moreover, the diurnal temperature cycle during the Neoproterozoic was shorter and probably less effective as a geocryologic agent than the present cycle, given the late Neoproterozoic 21.9 hour length of day [*Williams*, 2000]. Hence we affirm our conclusion, based on much geomorphologic and meteorologic data, that large-scale periglacial sand wedges are best ascribed to large seasonal changes of temperature.

3.3. Strong Winds

The occurrence of extensive periglacial eolianites in northwestern Africa [*Deynoux et al.*, 1989] and southern Australia [*Williams*, 1998] and glacial loessites in Norway and Spitsbergen [*Edwards*, 1979] indicates a very windy Neoproterozoic glacial environment.

The Whyalla Sandstone in South Australia covers 25,000 km^2 in outcrop and subcrop and is coeval with the Elatina Formation of the Marinoan glaciogenic succession [*Preiss*, 1987; *Drexel et al.*, 1993; *Williams*, 1998]. The flat-lying formation is up to 165 m thick and comprises mainly medium- to very coarse-grained, well rounded, commonly bimodal quartzose sandstone that shows regional south-southeastward fining. Low-angle strata form the principal stratification type. Cross-bed sets up to 7 m thick (Figure 1d) occur mainly in the central area. A suite of large-scale periglacial structures in the basal 5 m of the formation includes two generations of sand wedges. Sedimentary features of the Whyalla Sandstone accord with an eolian sand sheet environment and the periglacial structures with a cold, arid, strongly seasonal climate. The attitude of cross-bedding, the direction of regional fining, and paleolatitudes determined for coeval Marinoan strata imply the regional movement of paleo-northwesterly winds obliquely toward or across the paleoequator [*Williams*, 1998], employing the geographic polarity currently favored for Neoproterozoic Australia [*Li and Powell*, 2001].

4. RECONCILING PALEOMAGNETIC DATA AND THE GEOLOGIC RECORD

How can equatorial paleolatitudes for Neoproterozoic glaciation be reconciled with extensive and long-lived open seas and strong seasonality near the paleoequator?

4.1. Large Non-dipole Components of the Proterozoic Geomagnetic Field?

The geocentric axial dipole (GAD) hypothesis underpins the whole paleomagnetic edifice although almost certainly it is only approximate. A GAD is a hypothetical ideal that assumes that after secular variation is averaged the geomagnetic field can be considered as a point dipole at the Earth's center aligned with the spin-axis. Because it is generally thought that the field originates from self-sustaining dynamo action within the liquid outer core, that is, not at the geometrical center of the Earth, it should not surprise that the GAD is an approximation. The dipole equation states that $\tan[I] = 2\tan[\lambda]$, where I is magnetic inclination and λ is paleolatitude. The elegant simplicity of the dipole equation probably has conferred upon it a degree of reverence that has deterred some from doubting its veracity, let alone seeking an improvement. However, the closer paleomagnetists look at the ever-growing database, the more apparent become small anomalies and departures from expectations.

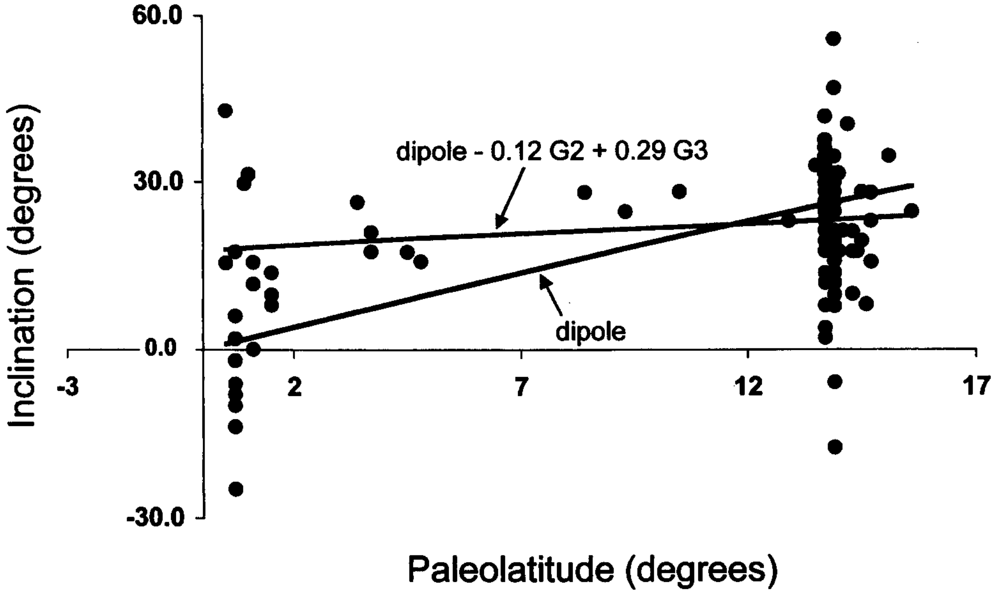

Figure 4. Extant paleomagnetic data for the Mackenzie and Sudbury dike swarms, and the best-fit (axial) dipole, quadrupole (G2) and octupole (G3) model compared to the pure axial dipole [*Schmidt*, 1999]. Sources of data: *Larochelle* [1967], *Fahrig and Jones* [1969], *Robertson* [1969], *Irving et al.* [1972], and *Park* [1974].

The geomagnetic field for the late Tertiary, a time for which there is an order of magnitude more data than for any other time, contained a persistent ~4% (zonal) quadrupole field. *Wilson and Ade-Hall* [1970] showed that poles from young rocks of Europe and Asia were 'far-sided', that is, they overshot the present north geographic pole by a few degrees. Numerous studies have demonstrated that it is a worldwide phenomenon [*Merrill et al.*, 1996]. *Merrill and McElhinny* [1977] suggested that regional average poles should be calculated in two ways, using the standard dipole relationship and using a modified relationship to correct for the quadrupole. So, how good is the GAD approximation for earlier geomagnetic times?

Evans [1976] examined the distribution of inclinations obtained from pole lists for rocks dating back to 600 Ma, and used absolute inclination, $|I|$, to combine data of both polarities. Making the assumption that the geographic locations where the sampled rocks had acquired their magnetizations were uniformly distributed on the globe, he compared the distribution of $|I|$ with that expected from a dipole. *Evans* [1976] concluded that there was no reason to doubt the validity of the GAD hypothesis since 600 Ma.

Piper and Grant [1989] applied this procedure for rocks dating back to 3000 Ma and also inferred the validity of the GAD hypothesis. However, *Kent and Smethurst* [1998] pointed out that *Piper and Grant* [1989] did not 'bin' their data according to geographic region and hence their results may be biased by an uneven regional spread of data. *Kent and Smethurst* [1998] analyzed the distribution of $|I|$ for Precambrian and Phanerozoic data (Holocene to 3500 Ma), binning the data temporally and geographically and examining separately data for crystalline and sedimentary rocks. They found the GAD hypothesis entirely adequate for Cenozoic and Mesozoic results, but their analysis suggested the persistence of small but significant non-dipole fields for the Paleozoic and the Precambrian: the distributions of $|I|$ for pre-Mesozoic results were consistent with a 10% quadrupole (G2 = 0.1) and a 25% octupole (G3 = 0.25). The magnitude of the octupole is surprising because the present octupole component is smaller than the quadrupole and it is generally assumed that the higher the order, the more quickly a multipole decays with distance from its source. *Kent and Smethurst* [1998] discussed other possible explanations for the Paleozoic and Precambrian data, including a biased distribution of continents.

If a supercontinent is defined with sufficient accuracy, the morphology of the field can be determined directly. Using this approach, *Van der Voo and Torsvik* [2001] showed that an axial octupole field eliminates the need for non-Bullard type Pangeas [cf. *Bullard et al.*, 1965]. However, their conclusions were challenged by *Muttoni et al.* [2003], who used a paleomagnetic database for igneous and volcanic rocks which are not susceptible to inclination shallowing. To many, nonetheless, 'Occam's razor' would dictate a small, realistic modification to the GAD rather than controversial tectonic solutions for continental misfits.

Another approach that, in principle, could constrain the non-dipole content of the geomagnetic field is to analyze paleomagnetic data for large dike swarms, provided the swarm is appropriately oriented and spans a sufficient latitudinal range. *Schmidt* [1999] analyzed paleoinclinations of the Mesoproterozoic Mackenzie and Sudbury dike swarms in Canada, using a modification of the methods developed by *Creer et al.* [1973] and *Georgi* [1974]. Although the ages of the Mackenzie dikes (1267 Ma) and the Sudbury dikes (1235 Ma) differ slightly [*LeCheminant and Heaman*, 1989; *Dudas et al.*, 1994], the proximity of their pole positions suggests little polar wander between times of intrusion. *Schmidt* [1999] was primarily interested in detecting long-term non-dipole features, over and above that averaged as secular variation. He also required data from a broad band of paleolatitudes and therefore combined data from both swarms. Polarity notwithstanding, the quadrupole and octupole components present in the dike data appear to be very similar to those proposed by *Kent and Smethurst* [1998]: the best fitting geomagnetic field for 1235–1267 Ma comprised a negative 10% zonal quadrupole (G2 = –0.01) and a positive 30% zonal octupole (G3 = 0.3) (Figure 4). Owing to symmetry, the sign of the quadrupole component is not apparent in the Kent–Smethurst analysis. However, the data for the Mackenzie and Sudbury dikes are from the 1960s and 1970s and do not meet present-day paleomagnetic standards. New, detailed paleomagnetic investigations of these dike swarms are required.

A Monte-Carlo simulation (Figure 5) was performed to ascertain the precision that such studies would require to confidently extract the non-dipole components from paleomagnetic data. Five transects along paleolatitudes from 5°S to 15°N in increments of 5° were modeled with magnetization directions consistent with a geomagnetic field containing zonal non-dipole components of G2 = –0.1 and G3 = 0.3. Mean paleomagnetic poles and precision parameters (K) were calculated for different values of G2 and G3. The most likely configuration of the geomagnetic field is assumed to be that associated with maximum K.

Means and standard deviations for G2 and G3 were calculated after 100 simulations for increasing errors in pole positions. It is apparent that for errors exceeding 3° or 4° the estimates of G2 and G3 may be very unreliable. This simple error analysis dictates that any new paleomagnetic investigation of the Mackenzie and Sudbury dikes should aim to yield paleomagnetic pole positions with A_{95} (semi-angle of cone of 95% confidence about mean pole) <4° and preferably <3°. Such precision, although exacting, is not unrealistic for these ideal paleomagnetic recorders.

Octupole and quadrupole contributions will affect paleolatitude determinations differently. A small (<30%) axial octupole field will not affect equatorial (and polar) paleolatitude interpretations, but an extreme (>50%) octupole content of the same sign as the dipole field produces a multi-valued inclination–latitude relationship with the field having zero inclination at ± 25° latitude as well as at the equator. A moderate quadrupole field will offset equatorial paleolatitudes; for example, a 30% quadrupole yields a 13° paleolatitude error. However, analyses of the geomagnetic field configuration using $|I|$ distributions cannot resolve a quadrupole field, hence any quadrupole content may be difficult to recognize in gross statistical analyses.

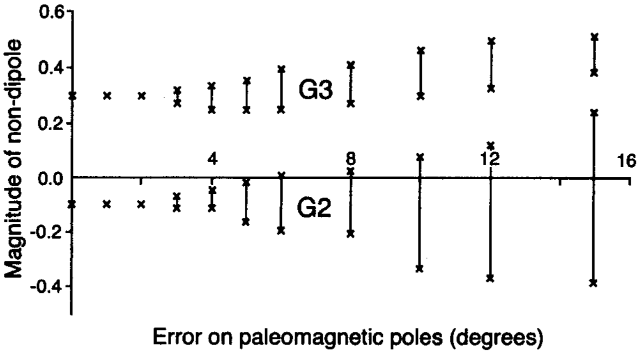

Figure 5. Results of Monte-Carlo simulations of a hypothetical paleomagnetic study of the Mackenzie and Sudbury dike swarms comprising five detailed traverses. Pole positions were calculated for each traverse assuming G2 = –0.1 and G3 = 0.3. Random errors were introduced to gauge the effect of noise apparent by the increase in the standard deviation of 100 trials. The bands plotted represent ± 1 σ.

The veracity of paleolatitudes may be tested through paleoclimate indicators. Large-scale periglacial wedges of Pleistocene age occur at 42.5°N in North America [*Wayne*, 1990] and at 41°S in South America [*Galloway*, 1965]. Hence a discrepancy of ~30° between paleomagnetic and true latitudes may resolve the paradox of a strongly seasonal climate, as implied by the occurrence of periglacial sand-wedge polygons, near the Marinoan paleoequator. An extremely large (50%) octupole component would be required to cause such major reinterpretation of the Marinoan paleomagnetic data, whereas two independent analyses suggest a Proterozoic octupole component no greater than 30%. Low paleolatitudes inferred for some Proterozoic glaciations thus remain a first-order geologic and geophysical enigma.

4.2. A Neoproterozoic Snowball Earth?

According to the snowball Earth hypothesis [*Kirschvink*, 1992; *Hoffman and Schrag*, 2002], negative $\delta^{13}C$ values for carbonates bracketing Neoproterozoic glacial deposits in Namibia reflect a collapse of biological activity in the surface of the world ocean, caused by a frozen-over ('snowball')

Earth achieved through a runaway ice-albedo feedback. It was argued that the Neoproterozoic world ocean was frozen to an average depth of >1 km and the mean global temperature was –50°C, with mean surface temperatures of –80°C to –110°C in high latitudes [*Baum and Crowley*, 2001]. The global freeze-over and resulting virtual shut-down of the hydrologic cycle lasted for up to 30 million years. In the absence of CO_2 sinks such as silicate weathering, volcanic outgassing during glaciation raised atmospheric CO_2 to 350 times the modern level. The ensuing extreme greenhouse conditions raised the mean global temperature to 40°C and abruptly (years to decades) ended the snowball state, leading to the deposition of 'cap' carbonates on glacial deposits. Soluble ferrous iron that accumulated in anoxic seawater during the global freeze-over was oxidized by the atmosphere upon melting of the ice cover, causing the precipitation of iron-formations at the *close* of glaciation. As discussed below, this snowball Earth scenario is difficult to reconcile with numerous, diverse geologic observations and interpretations.

A variety of geologic findings implies open seas during Neoproterozoic glaciations (see Section 3.1). These open seas were much more extensive and enduring than polynyas [cf. *Hoffman and Schrag*, 2002], which are only localized and transient openings between ice floes. Significantly, a runaway ice-albedo feedback is not exhibited by coupled ocean–atmosphere general circulation models for the Neoproterozoic, which imply that the Neoproterozoic sea-ice margin could not have advanced equatorward of 45° latitude [*Poulsen et al.*, 2001, 2002; *Poulsen*, 2003]. Floating marine ice, however, may have reached low latitudes [*Goodman and Pierrehumbert*, 2003].

Sellers [1990] found that the little precipitation and sublimation for a frozen-over Earth suppresses seasonal variation. Hence large seasonal changes of temperature accompanying Neoproterozoic low-paleolatitude glaciation (see Section 3.2) appear in conflict both with a snowball Earth and the near-snowball or 'slushball' Earth advocated by *Crowley et al.* [2001]. Proponents of a snowball Earth accordingly have advanced other, albeit speculative, explanations for Neoproterozoic sand wedges. *Hoffman* [2001] suggested that glacial surge cycles 10–100 years in duration cause the temperature changes required to produce periglacial wedge structures such as those *actively forming* in Antarctica and fossil wedges in South Australia. This idea is refuted by more than a century of observation and research on periglacial geomorphology and processes. *Hoffman and Schrag* [2002] then argued that seasonal temperature changes would be enhanced globally for a snowball Earth because of the low heat capacity of the solid, ice-covered surface and weak lateral heat transfer. Their argument fails because of the compelling evidence that during Neoproterozoic glaciations the seas were unfrozen across wide areas and for lengthy time-intervals and that many continental regions had patchy ice cover or were ice-free [*Deynoux*, 1982; *Williams*, 1998; *McMechan*, 2000]. As discussed in Section 3.2, the claim that diurnal temperature fluctuations could have produced large-scale periglacial polygonal wedges near the paleoequator [*Maloof et al.*, 2002] is refuted by the lack of such structures of Pleistocene age at high elevations on the equator.

Additional geologic observations conflict with the snowball Earth hypothesis.

Neoproterozoic glaciogenic successions commonly record numerous glacial advances and meltings, which are difficult to accommodate in the snowball Earth scenario [*Williams and Schmidt*, 2000; *Leather et al.*, 2002].

U–Pb zircon geochronology of ash beds below, within and above the glacial Gaskiers Formation in Newfoundland indicates that the glacial deposits are 580 Ma and formed in less than a million years [*Bowring et al.*, 2003]. The brevity of the Gaskiers glaciation is difficult to reconcile with requirements of the snowball Earth hypothesis.

Importantly, CO_2 sinks are not precluded during glacial times. CO_2 sublimes at –78.5°C at atmospheric pressure, hence the *mean* surface temperatures as low as –110°C modeled for Neoproterozoic global glaciation [*Baum and Crowley*, 2001] would have caused atmospheric CO_2 to precipitate and accumulate as the solid at high latitudes. In addition, carbonates are forming today below a permanent ice cover in saline lakes in Antarctica [*Walter and Bauld*, 1983]. Carbonate deposition therefore could have occurred during Neoproterozoic glaciations not only in the extensive unfrozen seas but also where the ice cover was thin.

The Rapitan iron-formation in Canada is succeeded by up to 600 m of tillite [*Yeo*, 1981] and the Sturtian iron-formation in South Australia is overlain disconformably, or with low-angle unconformity, by ~2700 m of mudstone containing dropstones and diamictite lenses [*Drexel et al.*, 1993]. These stratigraphic relationships show that, contrary to a prediction of the snowball Earth hypothesis, glaciation continued well after the deposition of iron-formation had ceased. Geologic and geochemical data are consistent with Neoproterozoic iron-formations in Canada, Alaska, China, South Africa, Australia and Brazil having accumulated in hydrothermally influenced rift-basins [*Yeo*, 1981; *Young*, 1982, 1988; *Jiafu et al.*, 1987; *Breitkopf*, 1988; *Neale*, 1993; *Trompette et al.*, 1998]. Volcanism was associated with the deposition of iron-formation in Canada [*Yeo*, 1981], Alaska [*Young*, 1982], China [*Jiafu et al.*, 1987] and Namibia [*Breitkopf*, 1988] and Sturtian glacial rocks in Australia [*Preiss*, 1987], supporting a hydrothermal origin. According to *Young* [2002], such iron-formations were precipitated in Red Sea rift-type basins when Fe-charged brines were mixed with 'normal' seawater through glacially driven thermal overturn.

Marinoan glaciation caused no major change in the nature of Vendian acritarch populations in Australia [*Grey et al.*, 2003]. Radical palynofloral change did not occur until about 15 million years after the end of Marinoan glaciation and coincides with the actual or estimated position of the ~580 Ma Acraman impact ejecta horizon [*Grey et al.*, 2003; *Williams and Wallace*, 2003]. Although the snowball Earth hypothesis ascribes late Neoproterozoic biotic changes to the aftermath of severe glaciation, *Grey et al.* [2003, p. 459] found that 'the predicted scenario is not supported by analysis of postglacial acritarch distributions.' Furthermore, microbial fossils in a carbonate unit associated with glacial deposits of the Neoproterozoic Kingston Peak Formation in California show that microbial life and trophic complexity survived low-latitude glaciation [*Corsetti et al.*, 2003]. The micropaleontologic record does not support catastrophic perturbation of the marine biosphere during postulated snowball Earth times.

Because numerous observations conflict with the basic tenets and predictions of the snowball Earth hypothesis, we conclude that this hypothesis cannot adequately reconcile the Neoproterozoic paleomagnetic and geologic records.

4.3. A Proterozoic High Obliquity?

The obliquity of the ecliptic (ε) has a value of 23.3 ± 1.3° and controls the seasonal cycle and climatic zonation. A Proterozoic high obliquity per se would not have been a primary trigger for glaciation, but during glacial intervals would have influenced the latitudinal distribution of glaciation and the global environment, as follows:

(1) The ratio of solar radiation received annually at either pole to that received at the equator increases with increase in obliquity, and latitudes $\leq 40°$ receive less solar radiation annually than high latitudes for $\varepsilon > 54°$ [*Williams*, 1975, 1993; *Oglesby and Ogg*, 1998; *Jenkins*, 2000, 2001, 2003; *Donnadieu et al.*, 2002]. Modeling with an obliquity between 60–90°, a 6% reduction of the solar constant and favorably located continents showed that ice sheets can form in latitudes $\leq 40°$ [*Donnadieu et al.*, 2002]. Ice sheets can spread to higher latitudes depending on the value of obliquity and the paleogeographic configuration.

(2) The amplitude of the global seasonal cycle increases with increase in obliquity [*Williams*, 1975, 1993]. For a high obliquity, all latitudes would experience strong seasonality and the mean monthly temperature range at the equator could reach 40°C [*Donnadieu et al.*, 2002].

(3) Zonal surface winds such as the tropical easterlies and mid-latitude westerlies reverse for $\varepsilon > 54°$ as the circulation in 'Hadley cells' reverses [*Hunt*, 1982]. In addition, the monotonic temperature gradient directed from the summer to the winter pole [*Hunt*, 1982] would cause atmospheric circulation across the equator around solstices.

(4) Climatic zonation is relatively weak for $\varepsilon > 54°$. Hence, as noted by *Williams* [1975, 1993], latitude-dependent climates would be unstable and any cyclic or chaotic orbital variations may cause abrupt and extensive climate changes and the stratigraphic juxtaposing of cold- and warm-climate deposits.

Features of Neoproterozoic glaciogenic successions and glaciations that accord with a high obliquity include:

- Glaciomarine deposition in low paleolatitudes.
- Large seasonal changes of temperature (~40°C mean monthly range) together with extensive, long-lasting open seas and ice-free continental regions near the paleoequator.
- Strong paleo-northwesterly winds in low paleolatitudes directed obliquely toward or across the paleoequator.
- The close association of glacial deposits and carbonates of apparent warm-water origin.
- The relatively powerful semiannual and annual signals revealed by Fourier spectral analysis of the 60-year Elatina rhythmite record [*Williams*, 1993].

The high-obliquity hypothesis must include mechanisms to give $\varepsilon > 54°$ during the pre-Ediacarian and then reduce the obliquity to $< 54°$ during the ~130 million years prior to Late Ordovician circum-polar glaciation. A corollary of the latter requirement is that the accompanying reduction of global seasonality triggered or accelerated the evolutionary radiation in the Ediacarian and Cambrian [*Williams*, 1993].

The widely accepted 'single giant impact' hypothesis for the Moon's origin [*Taylor*, 1987; *Canup and Asphaug*, 2001] requires a *glancing* impact of a body at least the size of Mars with the proto-Earth (impactor:target-planet mass ratio of ≥ 0.1), which would likely result in an obliquity of ~70° or more (Figure 6). *Dones and Tremaine* [1993] confirmed that the obliquity of a planet resulting from impacts by a few large bodies 'is likely to be substantial', with a probability of 0.73–0.91 that a single giant impact would have induced an obliquity of >24° for the early Earth. *Laskar et al.* [1993] found that a high obliquity would vary chaotically between 60° and 90° on a million-year time-scale, and the Precambrian (pre-Ediacarian) obliquity may have varied between those values [*Williams*, 1997]. The concept of a Precambrian high obliquity finds support in modeling of the Archean and Proterozoic climate [*Jenkins*, 2000, 2001, 2003]. Hence the long-held belief that the early Earth's obliquity was <23° [*Goldreich*, 1966] must be queried.

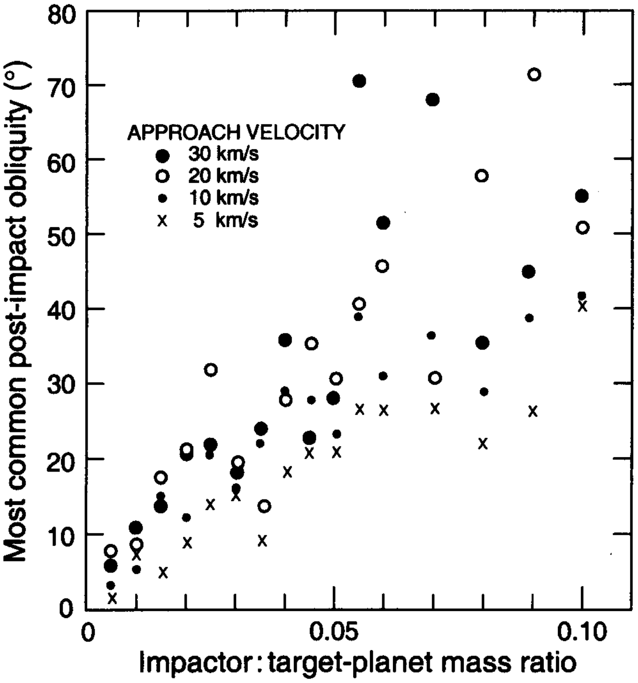

Figure 6. Impact-induced obliquity of an Earth-like target planet for a range of approach velocities and impactor:target-planet mass ratios. The target planet has a pre-impact obliquity of 0°. Each point is the most frequent value in a run of 500 impacts at a specified impactor mass. Adapted from *Hartmann and Vail* [1986].

Collisions between the Earth and impactors subsequent to the formation of the Moon are most unlikely to have caused significant change in the Earth's obliquity. An impactor several times the size of the envisaged K–T boundary bolide, for example, would alter the obliquity by only 0° 00′ 02″ under optimum conditions of impact [*Dachille*, 1963]. Moreover, the dynamical effects of terrestrial impact over time would have been random and would not have led to cumulative change in obliquity.

Slow change in obliquity, however, may have resulted from the action of luni-solar torques during Earth history. The mechanism of 'obliquity–oblateness feedback' or 'climate friction' for change in obliquity of the Earth and Mars involves changes in global mass distribution and the rate of spin-axis precession, with feedback loop, due to the waxing and waning of ice sheets [*Bills*, 1994; *Rubincam*, 1995, 1999; *Ito et al.*, 1995]. In principle, either secular increase or decrease of the Earth's obliquity may result, depending on the rate of ice-sheet variation and the rate of solid-earth deformation, which is controlled by mantle viscosity. *D. M. Williams et al.* [1998] suggested this mechanism as a possible cause of the postulated obliquity-decrease since the Neoproterozoic. Their findings were refuted by *Levrard and Laskar* [2003], and no study to date has been able to explain such a relatively rapid decrease in obliquity. As stated by *Rubincam et al.* [1998], however, 'Unlike tidal friction, a topic two centuries old, climate friction is such a new field that we can't yet determine its importance for changing our planet's tilt. Only time will tell.' The effects of late Neoproterozoic to early Paleozoic superplumes associated with the breakup of Rodinia [e.g., *Williams*, 1994b; *Puffer*, 2003] on global mass distribution and obliquity–oblateness feedback also merit examination.

5. CONCLUSIONS AND FUTURE DIRECTIONS

Non-GAD, snowball Earth, slushball Earth, or high obliquity? All these hypotheses offer explanations for Neoproterozoic low-paleolatitude glaciation but none is established firmly. We emphasize the following points:

(1) The snowball and slushball Earth and high obliquity hypotheses assume that the Proterozoic geomagnetic field approximated a GAD.

(2) A discrepancy of ~30° between paleomagnetic and true latitudes may account for low-paleolatitude glaciation, but this requires extreme (>50%) octupole content of the same sign as the dipole field. Analyses of Proterozoic paleomagnetic data suggest the possibility of an octupole component but no greater than 30%. Much remains to be done in this area.

(3) Geologic observations presented here indicating extensive and long-lived open seas and strong seasonality near the paleoequator, and additional data, conflict with the snowball Earth hypothesis.

(4) Demonstration of a reliable high paleolatitude for a pre-Ediacarian glaciomarine deposit would not confirm a snowball–slushball Earth because the paradox of glaciomarine deposition, extensive and long-lived open seas, and strong seasonality near the Neoproterozoic paleoequator would remain unresolved. A non-GAD or a high obliquity with favorable paleogeography may be preferred.

(5) If future work shows that no continent occupied high paleolatitudes during Neoproterozoic glaciations, non-GAD and high-obliquity hypotheses would remain viable because of the strong evidence for extensive open seas and marked seasonality near the paleoequator.

(6) Demonstration of diachronous glaciation in low paleolatitudes would militate against a snowball–slushball Earth and favor a high obliquity or a non-GAD. However, a finding of synchronous glaciation in low paleolatitudes would not necessarily favor a snowball–slushball Earth, because synchronous glaciation is possible with a high obliquity if glacial intervals saw little plate-motion.

The Neoproterozoic paleomagnetic and geologic records at present cannot be reconciled satisfactorily. Continued wide-ranging studies of Proterozoic glacial and nonglacial rocks and further investigations into the nature of the Proterozoic geomagnetic field and obliquity–oblateness feedback mechanisms are demanded.

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